AGU PUBLICATIONS

Geophysical Research Letters

RESEARCH LETTER
10.1002/2014GL062512

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- Last glacial AMOC changes preceded ITCZ shifts associated with Heinrich stadials

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Citation:
Burckel, P., C. Waelbroeck, J. M. Gherardi, S. Pichat, H. Arz, J. Lippold, T. Dokken, and F. Thil (2015), Atlantic Ocean circulation changes preceded millennial tropical South America rainfall events during the last glacial, *Geophys. Res. Lett.*, *42*, 411–418, doi:10.1002/2014GL062512.

Received 11 NOV 2014
Accepted 24 DEC 2014
Accepted article online 5 JAN 2015
Published online 16 JAN 2015

Atlantic Ocean circulation changes preceded millennial tropical South America rainfall events during the last glacial

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Abstract During the last glacial period, Greenland's climate shifted between cold (stadial) and warm (interstadial) phases that were accompanied by ocean circulation changes characterized by reduced Atlantic Meridional Overturning Circulation (AMOC) during stadials. Here we present new data from the western tropical Atlantic demonstrating that AMOC slowdowns preceded some of the large South American rainfall events that took place during stadials. Based on ^{231}Pa/^{230}Th and Ti/Ca measurements in the same sediment core, we determine that the AMOC started to slowdown 1420 ± 250 and 690 ± 180 (1σ) years before the onset of two large precipitation events associated with Heinrich stadials. Our results bring unprecedented evidence that AMOC changes could be at the origin of the large precipitation events observed in tropical South America during Heinrich stadials. In addition, we propose a mechanism explaining the differences in the extent and timing of AMOC slowdowns associated with shorter and longer stadials.

1. Introduction

Both paleoproxy records and numerical modeling indicate that the tropical and northern North Atlantic regions are tightly linked on centennial to millennial timescales. Sediment records from the Cariaco Basin show that tropical Atlantic Ocean and atmosphere changes correlate with variations in the North Atlantic region since the last glacial period [*Deplazes et al.*, 2013; *Hughen et al.*, 1996; *Lea et al.*, 2003; *Peterson et al.*, 2000]. Some modeling studies have suggested that these regions may be connected via a coupled atmosphere-surface ocean mechanism [*Chiang and Bitz*, 2005; *Chiang et al.*, 2003], whereas other numerical simulations using fully coupled atmosphere-ocean models have indicated that a reduction in the Atlantic Meridional Overturning Circulation (AMOC) flow rate could induce a southward displacement of the Intertropical Convergence Zone (ITCZ) [*Kageyama et al.*, 2009; *Menviel et al.*, 2014; *Swingedouw et al.*, 2009].

To improve our understanding of the linkage between ocean circulation changes and tropical climate, we studied sediment cores from the western tropical Atlantic. The western side of the tropical Atlantic is the locus of the western boundary currents and is therefore ideal to monitor changes in the flow rate of the water masses constituting the AMOC [*Rhein et al.*, 1995]. Moreover, this region is of major importance to understand the mechanisms behind millennial-scale tropical atmospheric changes. Indeed, during the last glacial, millennial variability in precipitation over tropical South America is recorded in various archives such as marine or lake sediments and speleothems [*Arz et al.*, 1998; *Cheng et al.*, 2006, 2013; *Cruz et al.*, 2009; *Deplazes et al.*, 2013; *Jaeschke et al.*, 2007; *Jennerjahn et al.*, 2004; *Wang et al.*, 2004, 2007]. Precipitation changes appear to be controlled by the latitudinal position of the ITCZ and appear to covary over South America in the 0–30°S latitude band east of the Andes.

Under modern conditions, the Brazilian northeastern region, called Nordeste (Figure 1), is a semiarid region. Periods of intense precipitation (March–April) in the Nordeste occur when the ITCZ is at its southernmost position and the interhemispheric tropical Atlantic sea surface temperatures (SSTs) gradient is the weakest [*Hastenrath*, 2011]. During boreal summer, the SST gradient steepens and the ITCZ moves northward, following the position of maximum SST above which convection occurs.

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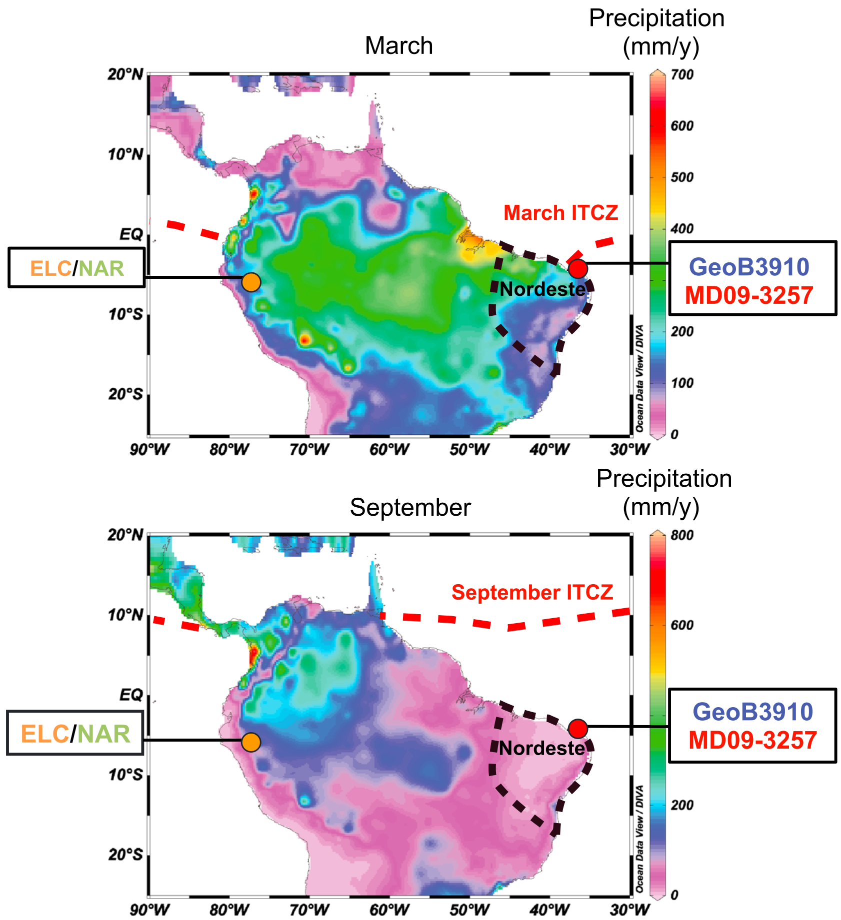

Figure 1. Map of the study area. The red circle indicates the location of the studied marine sediment cores, the orange circle indicates the position of the El Condor (ELC) and Diamante (NAR) caves [*Cheng et al.*, 2013], the black dashed line delineates the Nordeste region, and red dashed line shows the position of the present-day September and March ITCZ over the Atlantic and Pacific Ocean. The color scale indicates mean precipitation over South America (mm/yr) for the months of March and September averaged over the period 1950–1996 A.D. [*Willmott and Matsuura*, 2000].

2. Material and Methods

Cores MD09-3257 (04°14.69′S, 36°21.18′W, 2344 m water depth) and GeoB3910 (04°14.7′S, 36°20.7′W, 2362 m water depth) were retrieved from approximately the same location on the North Eastern Brazilian margin (Figure 1). At present, the cores site is bathed by the upper limb of the North Atlantic Deep Water, originating from the northern North Atlantic and circulating between 1200 and 2500 m depth with a transport of approximately 11 sverdrup [*Schott et al.*, 2003]. Numerous studies indicate that the northern sourced deep water flowed at shallower depth (above ~2000 m) during the last glacial compared to modern times [*Curry and Oppo*, 2005; *Duplessy et al.*, 1988; *Gherardi et al.*, 2009]. Hence, the location of the sediment cores studied here is ideal to monitor past changes in the AMOC upper circulation cell as well as their relation to millennial-scale precipitation events that are associated with the shifts in the ITCZ position.

During the last glacial, millennial-scale southward shifts of the ITCZ led to periods of increased rainfall in the Nordeste. Increased precipitation is marked by a decrease in $\delta^{18}O$ in South American speleothems [*Cheng et al.*, 2013; *Cruz et al.*, 2009; *Kanner et al.*, 2012; *Mosblech et al.*, 2012; *Wang et al.*, 2004, 2007] and increased sedimentary Ti/Ca ratios in marine sediment cores off the northeastern Brazilian coast [*Jaeschke et al.*, 2007] (Text S1 and Figures S1 and S2 in the supporting information). Precipitation events recorded in South American speleothems can be considered synchronous within age uncertainties (supporting information Text S2 and Figures S3–S5). These precipitation events were shown to take place during cold stadial periods recorded in Greenland ice cores [*Cheng et al.*, 2013; *Cruz et al.*, 2009, 2005; *Kanner et al.*, 2012; *Mosblech et al.*, 2012; *Wang et al.*, 2006, 2004]. In what follows, we call Heinrich stadials the six stadials characterized by the occurrence of

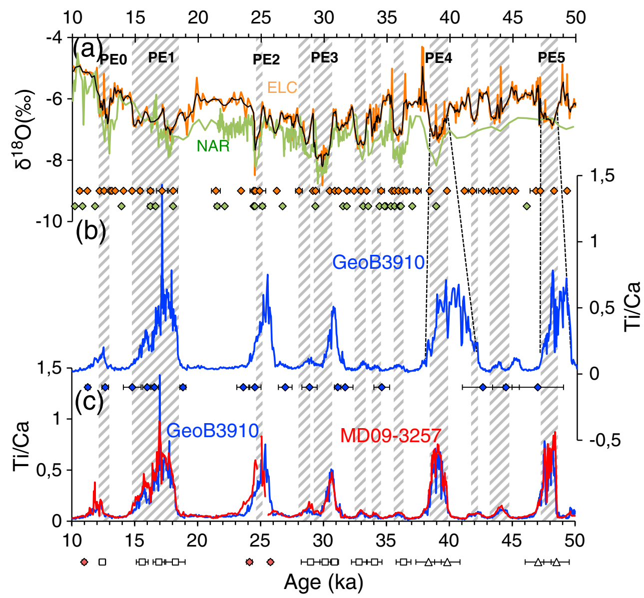

Figure 2. Comparison between (a) the U-Th dated δ^{18}O records of ELC/NAR speleothems [*Cheng et al*., 2013] and (b) the ^{14}C-dated Ti/Ca record of core GeoB3910 [*Jaeschke et al*., 2007]. (c) Ti/Ca record of marine sediment cores GeoB3910 [*Jaeschke et al*., 2007] and MD09-3257 on their final age models. Colored diamonds show the positions of the U-Th dates for speleothems and of the ^{14}C dates for marine sediment cores with their 2 standard deviation error bars. Diamonds below the *x* axis depict the ^{14}C dates used to establish the age model of core MD09-3257, empty squares show the position of the tie points used to establish the age model of core MD09-3257 by correlation of its Ti/Ca signal with that of core GeoB3910, and empty triangles are the speleothem tie points used to establish the age model of cores MD09-3257 and GeoB3910 beyond the ^{14}C dating range (see text and Tables S1 and S2). Striped bands delineate the precipitation events defined on the ELC speleothem δ^{18}O record (supporting information Text S2 and Figure S3).

Heinrich events, i.e., thick layers of ice-rafted detritus in middle- to high-latitude North Atlantic sediments, indicative of massive iceberg surges. We define PE0 as the precipitation event occurring during the Younger Dryas, and PE1 to PE5, those occurring during Heinrich stadials 1 to 5, respectively (supporting information Text S2 and Figure S3).

Precipitation events associated with Heinrich stadials (PE1–PE5) are reflected by sharp and large Ti/Ca peaks, whereas precipitation events associated with Dansgaard-Oeschger stadials are reflected by small Ti/Ca increases, likely indicative of less intense rainfall. Using ^{14}C dating, we could verify that all the precipitation events recorded in the studied marine cores are synchronous with precipitation events in speleothems for the 10–38 cal kyr B.P. (ka hereafter) period (Figure 2 and supporting information Text S4). This led us to synchronize to speleothems δ^{18}O records the two high-Ti/-Ca events, PE4 and PE5, which occur during the 38–50 ka period, i.e., beyond the range of precise ^{14}C dating. Doing so, we observe that the two small precipitation events seen in marine sediments between 38 and 50 ka are >synchronous with those seen in speleothems, which further confirms the synchronism between precipitation events recorded in marine sediments and speleothems. Importantly, the resulting marine cores age scales are based on radiometric dating, and hence completely independent from the Greenland ice core chronology (see supporting information Text S4 and Figure S6 for more information on the age model).

We used sedimentary ^{231}Pa/^{230}Th (Pa/Th hereafter) to reconstruct changes in the AMOC flow rate, with higher sedimentary Pa/Th corresponding to slower flow rates, and conversely [*Gherardi et al*., 2009; *Guihou et al*., 2010; *McManus et al*., 2004; *Yu et al*., 1996] (supporting information Text S5). Although a recent study questioned the capacity of dissolved Pa/Th to reflect the flow rate of water masses in the South Atlantic [*Deng et al*., 2014], sedimentary Pa/Th in the western tropical Atlantic is at present mainly controlled by oceanic circulation [*Lippold et al*., 2011]. While caveats may apply to this proxy in regions of high particle flux, we could verify that during the last glacial, the main drivers of the Pa/Th variability in our core remained the changes in oceanic circulation intensity (supporting information Text S5 and Figure S7). Sedimentary Pa/Th measurements thus provide unique information on the dynamics of the water mass overlying the study site. This information is complementary with respect to that provided by the δ^{13}C of the benthic foraminifer *Cibicides wuellerstorfi* which reflects bottom water nutrient content and is classically interpreted in terms of bottom water ventilation (supporting information Text S5).

We focused our Pa/Th measurements on two periods centered on precipitation events PE2 and PE4 because they correspond to different background climates, with much lower sea levels during PE2 than during PE4 [*Lambeck and Chappell*, 2001]. No Pa/Th measurements were done within the high-Ti/-Ca values associated with PE2 as this event is affected by a turbidite (supporting information Text S1).

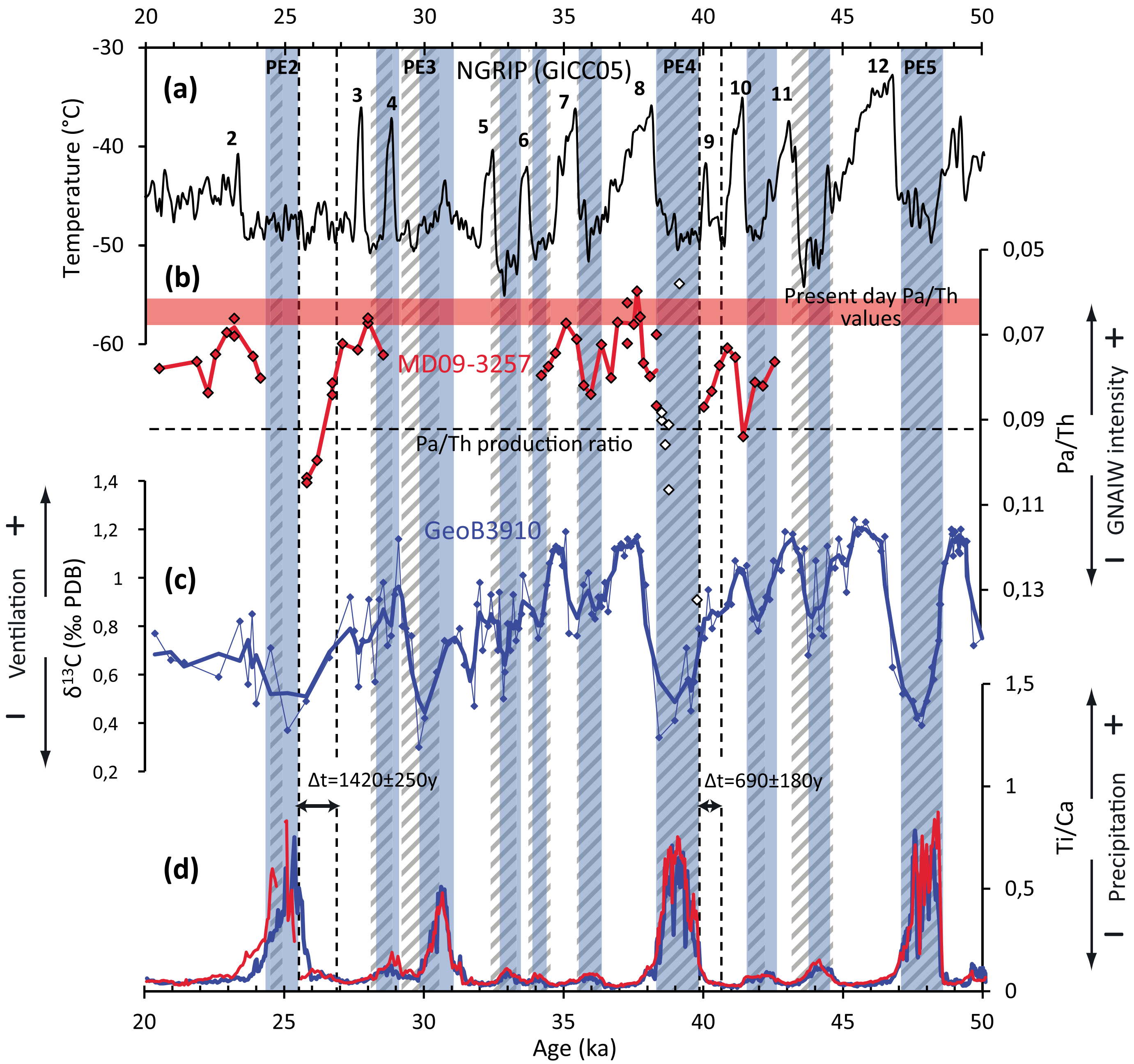

Figure 3. Comparison between (a) Greenland NGRIP temperature signal on the GICC05 timescale [*Kindler et al.*, 2014] and Brazil (b) MD09-3257 Pa/Th, (c) GeoB3910 *Cibicides wuellerstorfi* $\delta^{13}C$, and (d) GeoB3910 [*Jaeschke et al.*, 2007] (blue) and MD09-3257 (red) Ti/Ca records. Striped bands delineate the precipitation events defined on the ELC speleothem $\delta^{18}O$ record as in Figure 2. Blue bands delineate the precipitation events as seen in the MD09-3257 Ti/Ca record (see text). Vertical black dashed lines indicate the Ti/Ca increases and onset of the Pa/Th increases before PE2 and PE4 in core MD09-3257. In Figure 3a numbers indicate the DO events. In Figure 3b the red line connects average Pa/Th values when there are replicate measurements, the white symbols indicate Pa/Th values that should be interpreted with caution (see supporting information Text S5), and the red horizontal band indicates the present-day Pa/Th value (0.065 ± 0.003, supporting information Text S5). Error bars on Pa/Th measurements are given in the supporting information and Figure S8. No Pa/Th measurements could be made during PE2 because of the occurrence of two small turbidites (supporting information Text S1). In Figure 3c the thick dark blue line is the three points running average of the *Cibicides w.* $\delta^{13}C$ record.

3. Results

Comparing the Ti/Ca marine records with Greenland air temperature record on their respective independent calendar age scales, we observe that periods of increased precipitation over our study area take place during cold periods over Greenland (Figure 3). This is the case for both large precipitation

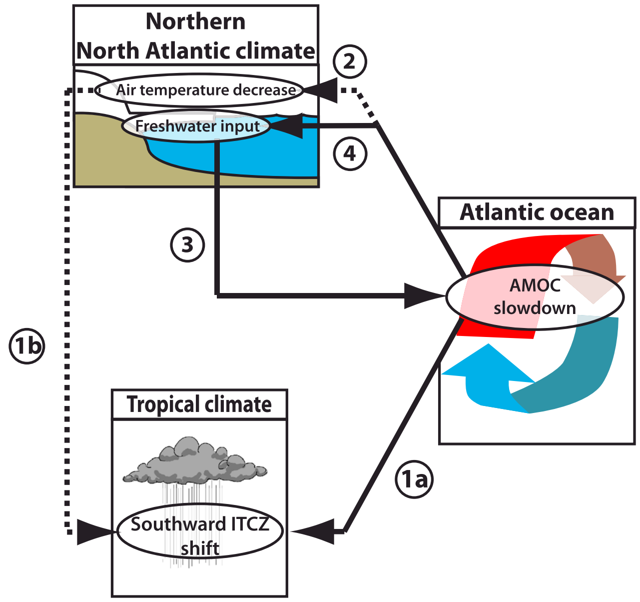

Figure 4. Illustration of the interactions between northern North Atlantic climate, tropical climate, and the AMOC. Black arrows illustrate causal mechanisms between triggers and climatic response discussed in the text. Dotted arrows indicate additional potential mechanisms (see text).

events PE2 to PE5 and smaller ones (Ti/Ca < 0.2) associated with the stadials preceding Dansgaard-Oeschger (DO) 5, 6, 7, 10, and 11. The only exception is the small precipitation event following PE3 in our sediment core that appears to coincide with the DO4 interstadial. However, the timing of this precipitation event can be reconciled with that of the stadial preceding DO3 within dating uncertainties.

Our measurements show that the Pa/Th signal rises from approximately modern day values prior to precipitation events PE2 and PE4, to values close to that of the production ratio (0.093) at their onset (Figure 3b). This indicates that the overlying water mass flow rate was strongly reduced at the onset of PE2 and PE4. Similarly, decreased flow rate accompanied the precipitation events of small amplitude related to DO stadials, as illustrated by the increase in Pa/Th values during each DO-related stadial for which we have the data. Moreover, increases in Pa/Th are synchronous with decreases in *C. wuellerstorfi* δ^{13}C (Figure 3c) that can be interpreted as reflecting reduced bottom water ventilation. The concurrent decrease in benthic δ^{13}C and increase in Pa/Th ratio in our cores thus indicates unambiguous decreases in the flow rate of the water mass occupying the first ~1000 m above the core site (i.e., the depth of acquisition of the Pa/Th signal [*Thomas et al.*, 2006]) during stadials. Considering the relatively high benthic δ^{13}C prevailing during DO interstadials, we interpret the Pa/Th increase and δ^{13}C decrease observed during stadials as a decrease in the flow rate of the northern sourced water mass between ~1300 and 2300 m, and hence of the AMOC upper circulation cell. Alternatively, a Pa/Th increase and δ^{13}C decrease could also be interpreted as a shoaling of the northern sourced water mass and presence of sluggish southern sourced waters above our core site. However, such a configuration of the AMOC would likely also imply a reduction in the flow rate of the AMOC upper circulation cell.

Sedimentary Pa/Th starts rising before the Ti/Ca increase associated with PE4 and PE2 in core MD09-3257 (Figure 3, see also supporting information Text S6 and Figure S9). Because Ti/Ca and Pa/Th values were measured in the same core, the leads are directly measurable in terms of core length: Pa/Th starts rising nearly 5 cm before the onset of PE4 and more than 9 cm before the onset of PE2. Converting these depth intervals into time intervals, the lead of the Pa/Th signal with respect to the Ti/Ca signal is estimated to be 690 ± 180 (1σ) years and 1420 ± 250 (1σ) years at the onsets of PE4 and PE2, respectively. The slowdown of the AMOC upper circulation cell therefore significantly precedes the large precipitation events observed in the 0–30°S latitudinal band in South America. While the sequence of events is straightforward for PE4 and PE2, the phasing between AMOC slowdowns and small precipitation events related to DO stadials is more difficult to assess. If an AMOC slowdown precedes small precipitation events associated with DO6 and DO7, then the time lag between the beginning of the AMOC slowdown and of the precipitation event is much smaller than for precipitation events associated with Heinrich stadials (Figure S9).

4. Discussion

Our data bring new constraints on the possible triggering mechanisms of the millennial precipitation events observed in the 0–30°S latitudinal band in South America during the last glacial. Our results demonstrate that precipitation events corresponding to Heinrich stadials over our study area occurred 500 to 1700 years after the beginning of a slowdown of the AMOC upper circulation cell. Thus, there could be a causal link between the AMOC slowdowns and ITCZ shifts.

Climate models indeed show that an AMOC slowdown causes a decrease in interhemispheric oceanic heat transport (see *Kageyama et al.* [2010] for a review), which results in negative SST anomalies in the northern tropics and positive SST anomalies in the southern tropics. This mechanism leads to the southward shift of the convection zone, which migrates toward the warmer hemisphere (arrow 1a in Figure 4). However, current coupled ocean-atmosphere climate models simulate a response of tropical precipitation to AMOC changes after only a few hundred years [e.g., *Kageyama et al*., 2009], a lag significantly shorter than that measured for PE2 and PE4.

Other modeling studies produced southward ITCZ shifts in response to extended high-latitude ice sheets or sea ice cover [*Chiang and Bitz*, 2005; *Chiang et al*., 2003], without any change in the AMOC. In these models, surface cooling in the North Atlantic and Nordic Seas induces the progression of low SST toward the ITCZ latitudes, leading to a southward displacement of the ITCZ after only a few years (arrow 1b in Figure 4). Interestingly, if Greenland air temperatures reflect temperatures above the high-latitude North Atlantic, the records in Figure 3 do not support this mechanism as sole cause for the observed precipitation events, since PE2 starts ~2 kyr after the beginning of Heinrich stadial 2 cold period above Greenland. However, Greenland air temperatures might not reflect temperatures above the high-latitude North Atlantic. Indeed, our records indicate that changes in oceanic northward heat transport alone do not explain the NGRIP temperature record. D09 interstadial warm event indeed occurred in NGRIP while the AMOC was slackening before PE4, indicating that in this case, Greenland temperatures increased although oceanic interhemispheric heat transport was decreasing. Furthermore, temperature above Greenland did not markedly decrease during the AMOC slowdown preceding PE2 (Figure 3). These observations indicate that there is no direct linear response of Greenland temperatures to AMOC changes. This is in line with modeling studies showing that feedbacks involving sea ice extent have a large impact on Greenland temperatures [*Li et al*., 2005; *Menviel et al*., 2014]. Sea ice through its insulation effect could indeed decouple middle- to high-latitude North Atlantic air temperatures from Greenland temperatures. Hence, it is difficult to infer the effect of sea ice and middle- to high-latitude North Atlantic cooling on the position of the ITCZ from Figure 3 data. However, we cannot exclude that an AMOC slowdown and concurrent decrease in northward heat transport impacted the northern North Atlantic climate (arrow 2 in Figure 4) and induced increased sea ice cover and/or ice sheet extension which could, in turn, have caused a southward shift of the ITCZ (arrow 1b in Figure 4).

The large AMOC slowdowns recorded before PE4 and PE2 could result from reduced deep water convection following freshwater input in the northern North Atlantic (arrow 3 in Figure 4). Different origins of such a freshwater flux have been suggested: melting of ice sheets during active modes of the AMOC [*Broecker et al*., 1990], melting of icebergs and ice shelves during Heinrich or precursory events [*Peck et al*., 2006; *Vidal et al*., 1997], and pulses of fresh water from a Hudson Bay lake [*Johnson and Lauritzen*, 1995]. Extended sea ice cover in zones of deep water formation could also have caused slowdowns of the AMOC upper circulation cell [*Duplessy et al*., 1980].

Our data indicate that Pa/Th changes associated with DO events are of lower amplitude and do not feature the same large lead with respect to the Ti/Ca signal as for the precipitation events associated with Heinrich stadials. Indeed, only one Pa/Th data point during the stadial preceding DO10 exhibits a high value, and no other DO stadial Pa/Th value is higher than the Pa/Th values recorded at the onset of Heinrich stadials. Smaller increases in the Ti/Ca signal and lower Pa/Th values may reflect a more northerly final position of the ITCZ and smaller slowdown of the AMOC upper circulation cell. Different mechanisms could thus be at the origin of large and small precipitation events. The absence of a clear lead of the AMOC slowdown with respect to small precipitation events could imply that rapid atmospheric processes associated with extended ice sheet or sea ice cover are directly responsible for the small precipitation events. However, as slowdowns of the AMOC upper circulation cell also accompany small precipitation events, the influence of the AMOC on ITCZ shifts occurring during DO stadials cannot be ruled out.

We suggest that a positive feedback loop played a key role and lies at the heart of the difference between DO and Heinrich stadials. Recent studies have suggested that an AMOC slowdown and decrease in deep water formation could lead to warming of North Atlantic subsurface waters, and hence erosion of the ice shelves and destabilization of the Laurentide ice sheet [*Alvarez-Solas et al*., 2013; *Mignot et al*., 2007]. This would result in iceberg surges in the North Atlantic (arrow 4 in Figure 4) and further decrease deep water formation and slowdown the AMOC. We suggest that the large lead of the AMOC slowdown with respect to the onset of PE2 and PE4 results from the action of this positive feedback on the AMOC. Hence, an initial AMOC slowdown might have been amplified leading to a progressive decrease in overturning circulation. The AMOC would have progressively slowed down as long as this positive feedback operated, pushing the ITCZ to an increasingly southerly location. The observed large lead of the AMOC slowdown with respect to South American rainfall events during Heinrich stadials could therefore be explained by the progressive slowdown of the AMOC, sustained by the afore described feedback. Salt accumulation in low latitudes surface waters likely put an end to the progressive AMOC slowdown when the quantity of salt became sufficient to counter the influence of meltwater inputs at higher latitudes and cause the resumption of the oceanic interhemispheric heat transport [*Broecker et al*., 1990; *Paillard and Labeyrie*, 1994; *Schmidt et al*., 2004]. The reactivation of the overturning circulation would then cause the ITCZ to move back to its original, more northerly, position. We propose that, in contrast to the progressive amplification of the AMOC slowdown that took place during Heinrich stadials, the AMOC slowdowns that occurred during the stadials preceding DO events did not trigger the aforedescribed positive feedback. As a result, both the decrease in AMOC and latitudinal shift in the ITCZ taking place during DO stadials remained limited.

It is noteworthy that the proposed mechanism is consistent with the observed higher ice-rafted detritus content of North Atlantic sediments during Heinrich stadials than during DO stadials. Moreover, recent data from Antarctic ice cores show that atmospheric CO_2 progressively increased during each Heinrich stadial but remained constant or slightly decreased during DO stadials [*Ahn and Brook*, 2014]. The fact that the atmospheric CO_2 trend is drastically different during DO stadials than during Heinrich stadials points to the existence of an additional feedback involving the carbon cycle during Heinrich stadials that does not operate during DO stadials, very likely related to the large changes in ocean circulation demonstrated by our data.

5. Conclusions

Our results demonstrate that the AMOC upper circulation cell experienced marked slowdowns that started ~500 to ~1700 years before the major tropical South American precipitation events that occurred during Heinrich stadials and corresponded to southward shifts of the ITCZ. Our data thereby provide the first evidence that changes in Atlantic circulation preceded changes in other climatic variables during rapid climate changes of the last glacial period. Moreover, our results shed light on major differences in the mechanisms leading to the observed southward ITCZ shifts associated with Heinrich stadials compared to those associated with DO stadials. Our data thus open new prospects for testing causal mechanisms of rapid climate changes using fully coupled Earth system models. More specifically, although current climate models simulate a rapid response of the tropical climate to AMOC changes [*Kageyama et al*., 2010, 2009], the present study suggests that the observed large lead of AMOC changes with respect to the tropical climate could be simulated by coupling these climate models with ice sheet models.

Acknowledgments
The data of this paper are available in the supporting information. Pa/Th measurements were funded by the CNRS/INSU LEFE project ACCENT. This work is a contribution to the RETRO project, a joint European Science Foundation (ESF)/EUROMARC, funded by Research Council of Norway (RCN), France (CNRS/INSU), Germany, and the Netherlands. Core MD09-3257 was collected on board R/V *Marion Dufresne* during RETRO Cruise III, supported by ESF EUROMARC project RETRO, IPEV, and ANR project ANR-09-BLAN-0347. We thank the IPEV team, crew members of R/V *Marion Dufresne*, and all scientists who participated in RETRO Cruise III. We are thankful to M. Roy-Barman and E. Pons-Branchu for expert advices on Pa/Th measurements on LSCE MC-ICP-MS. We acknowledge C. Moreau, J.-P. Dumoulin, and the UMS ARTEMIS for AMS ^{14}C dates. We are grateful to A. Govin, M. Kageyama, J.-C. Duplessy, E. Michel, and J. Jouzel for constructive comments on an earlier version of this article. This is LSCE contribution 5279.

The Editor thanks Alex Thomas and an anonymous reviewer for their assistance in evaluating this paper.

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